WIND FIELD OBSERVATIONS BY DOPPLER RADAR IN A NEW ENGLAND SNOWSTORM

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ABSTRACT

Fields of wind velocity, divergence, and reflectivity were derived from Doppler radar measurements in a New England storm over a 6-hr. period. A layer of easterlies near the surface increased in depth and speed during the period. Above this layer the wind veered with height to SW and W, and then backed to SW. Changes in wind direction with time in the layer between 4 and 6 km. indicated the passage of mesoscale short wave troughs. Layers of convergence and divergence, alternating with height, appeared to be related to the wind regions.

1. INTRODUCTION

It is known by radar meteorologists that continuous precipitation is seldom uniform. This is particularly true aloft where streamers or cells of relatively heavy precipitation are observed to be imbedded in lighter precipitation. This variability of continuous precipitation must be associated with mesoscale features of the wind system. Doppler radar has been used to observe some of these features. Harrold [1] computed the field of divergence over a period of a few hours and found layers of convergence and divergence with maximum values surprisingly high—about 2×10^{-3} sec.$^{-1}$ Maximum values by Caton [2] and Browning and Wexler [3] were an order of magnitude less than those of Harrold.

In this paper the synoptic aspects of a late winter snowstorm, observed by Doppler radar at Sudbury, Mass., on March 15, 1967, are analyzed. On the morning of the storm the temperature over Nantucket at 1.5 km. was 4° C. with a southerly wind. This condition generally indicates rain over southern New England. The wind fields derived from the Doppler observations indicate the changing conditions which caused the precipitation to be maintained as snow.

The theory of determination of the wind field from Doppler radar is first briefly reviewed, with emphasis on the techniques used here. Time-height charts of the wind velocity, reflectivity of the precipitation, divergence, and vertical velocity are then analyzed.

2. THEORY

The theory of the determination of the wind field by Doppler radar has been given in previous studies. In particular, Browning and Wexler [4] showed how the different properties of the wind field can be determined from a harmonic analysis of data taken at azimuth intervals of 10°.

As the antenna rotates in azimuth, a quasi-sinusoidal pattern of Doppler wind speed results. The Doppler speed for any given azimuth θ and angle of elevation α is given by:

$$v=u\cos\theta\cos\alpha-(w+w_f)\sin\alpha \tag{1}$$

where u is the horizontal wind speed, w the vertical motion of the air, and w_f the mean fall speed of the precipitation particles (both positive downward). As used here $\theta=0$ is taken as the downwind direction, so that v is positive when directed away from the radar. In the March 15, 1967, storm the radar did not measure the sign of v, so that two positive maxima in the upwind and downwind directions were obtained.

For determining properties of the wind field best results are obtained by using all points of the Doppler speed curve as a function of azimuth. However, fairly accurate estimates of the wind direction can be made by using the midpoint between directions where the Doppler velocities are zero. The term $u\cos\alpha$ is obtained as the average of the absolute values of the maximum Doppler velocities in the upwind and downwind directions.[1]

The divergence of the wind is given by:

$$\operatorname{div}\mathbf{u}=\frac{1}{\pi r\cos^2\alpha}\oint v\,d\theta+\frac{2w_f\tan\alpha}{r} \tag{2}$$

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[1] A test of the method was made for the storm of December 13, 1965, analyzed by Browning and Wexler [4] by harmonic analysis of data at 36 points. The results showed that for a 3-hr. period, comprising more than 100 observations, wind directions obtained by the two methods differed by at most a few degrees and wind speed by a few tenths of a meter per second.

where w is neglected compared to w_f. The first term on the right hand side may be evaluated by measuring the difference of the areas under the Doppler curves in the upwind and downwind directions. The second term may be evaluated for snow by assuming a constant fall speed of 1 m. sec.$^{-1}$ For angles of elevation under 30° the error resulting from this assumption is relatively small.

A rough estimate of divergence may be made simply by assuming symmetric sine curves with respect to the maximum Doppler speeds in the upwind and downwind directions and computing the difference in their areas. However, unsymmetric patterns often occur (see fig. 1), so that considerable error may be involved in this method.

The vertical motion of the atmosphere may be determined from the divergence with the formula

$$w=\frac{1}{\rho_1}\int_0^z \rho \operatorname{div} \mathbf{u}\, dz \qquad (3)$$

where ρ is the density of air and ρ_1 the density at height z. It is to be noted that the vertical velocity at level z involves an integration process from the lowest level. Hence, errors at low levels are compounded at high levels. This is especially unfortunate for the March 15, 1967, case because wind speeds were extremely low, between 1.5 and 2.5 km., making accurate measurements of divergence impossible. Hence, the derived vertical velocity field for the March 15, 1967, case may be subject to error.

An estimate of the existence of cross wind shear can be made from the relative spacings in azimuth of the upwind and downwind maxima in Doppler velocity. For a perfectly symmetric wind, or for shear along the wind, these maxima are 180° apart. Cross-wind shear causes the maxima to be less than 180° apart in the direction of higher wind speed, and greater than 180° in the direction of lesser speed. Theoretically, the actual spacing may be used to determine the magnitude of the cross-wind shear [5]. However, the broad maxima make accurate measurements of the directions difficult. The magnitudes of shear are best studied from the field of deformation.

3. OBSERVATIONAL AND ANALYSIS PROCEDURES

The Porcupine Doppler radar with wavelength 5.4 cm., beam width 1°, 2 μsec. pulse and peak power of 17 kw., is located at Sudbury, Mass. Observations are made by a VAD technique (velocity azimuth display) [6] in which the antenna is rotated at a fixed elevation, and after a complete rotation is then stepped up in 2° intervals starting at 2° until the top of the storm is reached. Doppler data are obtained practically simultaneously at 10 different ranges, equally spaced and fixed from 12 to 17.4 km. Only the Doppler data at 17.4 km. are analyzed here. Reflectivity is measured at an intermediate range of 14.4 km.

The raw Doppler frequency spectra recorded on magnetic tape were subsequently analyzed by a Doppler average speed estimator to provide an estimate of the median of a Doppler power spectrum. This instrument electronically centers the Doppler spectrum in a discriminator, such that equal power exists on either side. Unfortunately, at low speeds the discriminator does not operate efficiently,[2] and the minimum speeds that can be measured is about 2 m. sec.$^{-1}$ However, for some reason, particularly at low wind speed the VAD pattern often shows a minimum near 5 m. sec.$^{-1}$ The output of this "mean tracker" is recorded in analog form and examples are seen in figure 1.

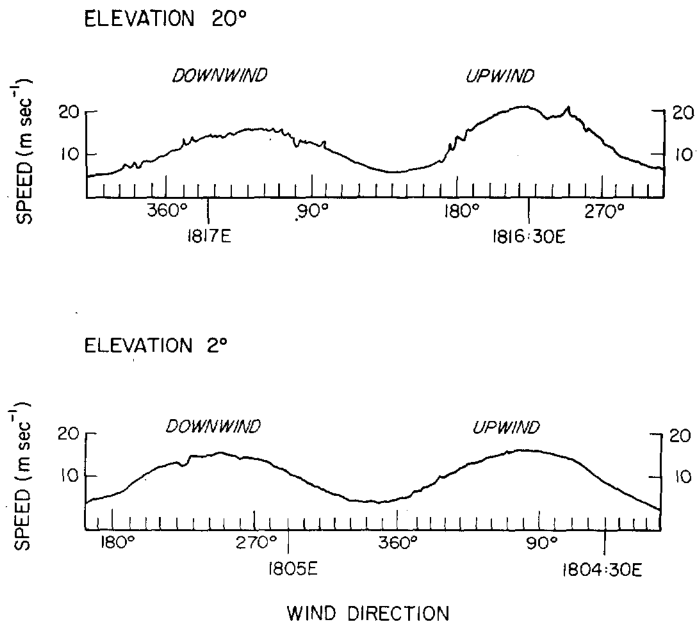

FIGURE 1.—VAD patterns at 2° elevation (0.7 km.) and at 20° (6.1 km.), March 15, 1967.

Wind speed is determined as the mean between the upwind and downwind maxima, while wind direction is obtained as the direction midway between minima. Although the Doppler velocity does not actually reach zero, the midpoint of the flat minimum in Doppler velocity is used (see fig. 1). When the wind speed is under 7 m. sec.$^{-1}$, the flat minimum covers too large an interval of azimuth, and in that case the actual directions of upwind and downwind maxima are used to obtain wind direction.

Divergence is determined by measuring the areas under the upwind and downwind portions of the VAD with a planimeter. The difference in area is then proportional to the divergence. Although portions of the curve with speeds less than 2 m. sec.$^{-1}$, and at times 5 m. sec.$^{-1}$, are missing, the areas beneath these portions are small for wind speeds exceeding 10 m. sec.$^{-1}$ For cases where the VAD curves are relatively flat due to low wind speed, areas are generally too small for accurate measurement. Nevertheless, even for those cases, the magnitudes of the "bulges" above the minimum can be used to indicate the existence of divergence or convergence. Some estimate of the magnitude of these differences were sometimes made, although un-

[2] By offsetting the zero speed level, low speeds can be measured with the mean tracker. However, this was not done for the March 15, 1967, storm.

doubtedly they were subject to considerable error. Cases of estimates of this kind are indicated by parentheses in the divergence diagram (fig. 5).

Figure 1 shows two examples of the VAD pattern, one at 2° the other at 20° elevation angle. At 2° both upwind and downwind portions of the VAD are symmetric. The Doppler minima are centered at 342° and 162° giving a wind direction of 72°. The upwind maximum is 16 m. sec.$^{-1}$ at 78° and the downwind 15.1 m. sec.$^{-1}$ at 250°. The area under the upwind curve is slightly greater than that under the downwind, indicating slight convergence. At 20° elevation the wind direction is 233°. The downwind direction of about 50° is in good agreement but upwind there is a secondary maximum at about 250° and a primary maximum at about 220° with a discontinuity in the VAD pattern at about 235°. Similar sharp discontinuities have been observed previously by Harrold [1] and by Lhermitte [7]. It is to be noted that despite the fact that the maximum in the upwind direction is 4 m. sec.$^{-1}$ higher than that in the downwind direction, the area under the upwind curve is only slightly higher than that under the downwind. With the fallspeed correction, a slight divergence is indicated. Thus, the difference in speed between upwind and downwind maxima does not necessarily provide a good estimate of the magnitude of convergence or divergence.

4. SYNOPTIC CONDITION MARCH 15, 1967

The surface weather map for 1300 EST, March 15, 1967, showed a primary center located over Virginia with a secondary located to the east just off the coast. Subsequently the systems combined and deepened and at 0100 EST, March 16, 1967, the center was located southeast of Nantucket. Snow was widespread over New England during the 12-hr. period. Aloft at 500 mb. at 1900 EST, March 15, 1967, there was a trough extending from Lake Ontario towards the SSE with almost a straight southwesterly flow of about 35 km. over New England. A close (post mortem) examination may indicate a short-wave trough to the east of Albany (ALB) since the winds there were coming from the WSW, while at Portland (PWM) and Nantucket (ACK) they were from the SW. The existence of a short-wave trough was better indicated at 700 mb. with the wind from 280° at New York (JFK) and 250° at ACK both at about 40 kt. Considerable cross wind shear existed to the north with the winds 10 kt. from 250° at ALB and 230° at PWM.

The lapse rate at PWM and ACK at 1900 EST followed the wet adiabat up to about 1 km. with an inversion just above. Conditions were then almost isothermal at about 4° C. at ACK and −9° C. at PWM to about 3.5 km., above which it was close to the wet adiabat.

5. WIND DIRECTION AND SPEED

Figures 2 and 3 show the wind direction and speed for the period 1315–2000 EST. There is a break in the data between 1700–1800 EST.

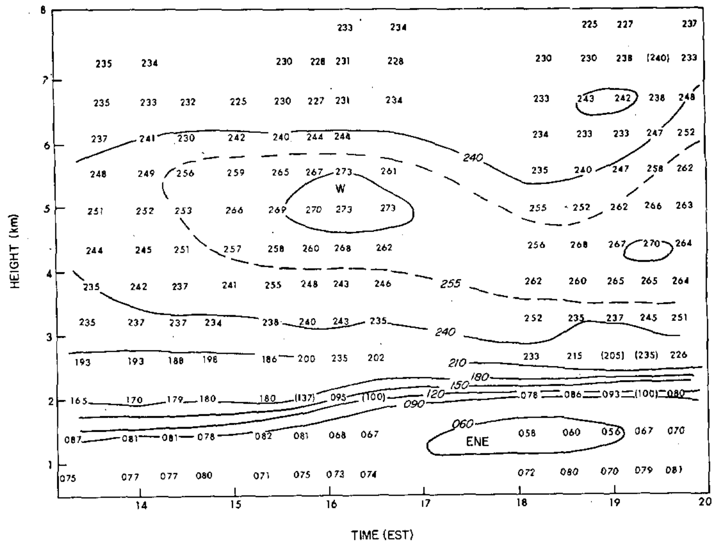

FIGURE 2.—Wind direction, March 15, 1967.

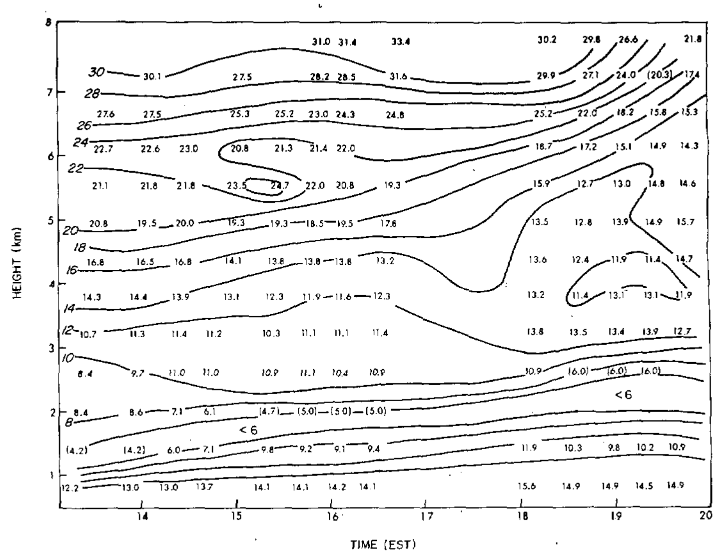

FIGURE 3.—Wind speed in m.sec.$^{-1}$, March 15, 1967.

At lower levels the wind direction is ENE and the speed increases from 12 to 15 m. sec.$^{-1}$ during the period. The depth of the easterly current is about 1.5 km. initially but it increases to about 2 km. after 1600 EST. The increasing depth of the easterly current may have been responsible for preventing warm advection and maintaining the precipitation as snow. The wind speed decreases with altitude in the easterly current and becomes very light (below 5 m. sec.$^{-1}$) in the middle of a transition layer where the wind shifts via the SW. The layer of SW current initially between 3 and 4 km. at about 12 m. sec.$^{-1}$ becomes compressed to a layer about 0.5 km. in depth by 2000 EST between the rising easterly current below and the descending westerlies just above and is accompanied by an increase in speed.

Centered at 5 km. is a flow initially from the WSW at about 20 m. sec.$^{-1}$, that turns more to the W with decreasing speed after 1500 EST. This westerly current later descends and is found near 4 km. at 1800 EST but then increases in depth and extends from about 3.5 to 6 km. by the end of the period.

Above the westerlies the wind shifts back to SW with speed increasing rapidly with height. The southwesterlies dip down temporarily to 5.5 km. at 1800 EST but then recede upward to above 6 km. During the 6½-hr. period, wind speeds in the SW flow decrease markedly: from about 27 to 15 m. sec.$^{-1}$ at 6.5 km.

The shift in the wind at 5.5 km. from WSW to W at the beginning of the period, then back to SW at 1800 EST and returning to W suggests the passage of mesoscale short-wave troughs.

At about 1500 EST there is a temporary increase of wind speed at 5.5 km., together with a temporary decrease at 6 km. This occurs just prior to the indicated passage of the mesoscale trough and is probably related to this feature. Another explanation is that it may be a temporary transfer of momentum from upper to lower levels in a region of strong wind shear. Similar breakdowns in strong wind shear have been observed by Boucher et al. [8] in a storm during which wind speeds were increasing with time.

6. FIELD OF REFLECTIVITY

Figure 4 shows the field of returned radar power in db.m. taken at a range of 14.4 km. This is proportional to the reflectivity, or the summation of the sixth powers of the diameters of the precipitation particles. Each reading represents an approximate average value over the 360° scan. Although the storm was continuous it was by no means steady in the different directions, and reflectivities sometimes varied by as much as 10 db.

Changes in reflectivity with height up to 4 km. are irregular. There are some periods where little or no change occurs, while occasionally the decrease up to 4 km. amounts almost to 10 db. Part of the decrease may be due to coalescence of ice crystals into snowflakes, which is favored at temperatures above −10° C. However, the rapid decrease in reflectivity aloft between 1330 and 1540 and the descent to near the surface of the heavier precipitation generated aloft early in the period is evidently the primary cause of the relatively large vertical gradient in reflectivity at about 1540.

Above 4 km. the wet adiabatic lapse favors precipitation growth, which is in agreement with the decrease of reflectivity with height above that level, generally by 4 to 5 db./km.

Since the reflectivity measurements represent an average around the perimeter of a circle 14.4 km. in radius, they cannot be directly compared with precipitation intensity measurements measured at the radar. The radar equation at a range of 14.4 km. is

$$P=1.37\times10^{-14}\,Z \tag{4}$$

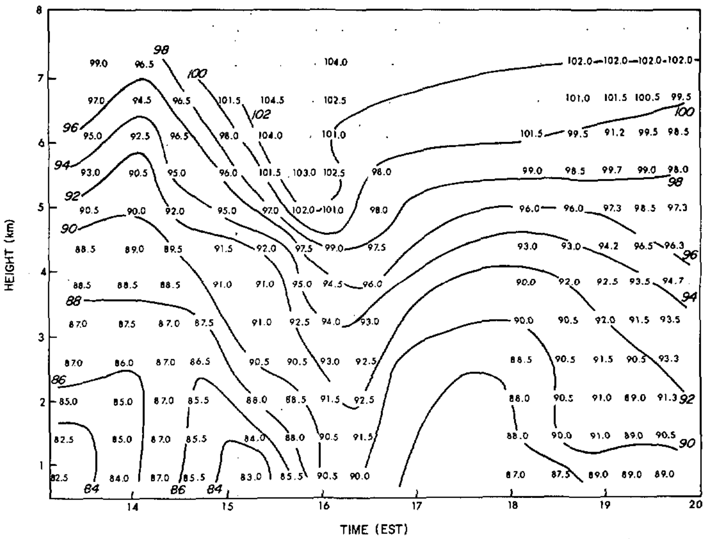

FIGURE 4.—Received power in db.m. from precipitation at a range of 14.4 km., March 15, 1967. For ice the reflectivity factor $Z=340$ mm.^{6}m.$^{-3}$ at 85 db.m. and 106 mm.^{6}m.$^{-3}$ at 90 db.m.

where P is the received power in watts and Z the reflectivity factor for ice in mm.^{6} m.$^{-3}$ Assuming an average P of 85 db.m. and inserting its equivalent 3.2×10^{-12} watts in the radar equation, we find $Z=234$ mm.^{6} m.$^{-3}$ For aggregate snow $Z=1000\ R^{1.6}$, approximately [9], so that $R=0.4$ mm. hr.$^{-1}$ The actual precipitation intensity during the period 1300–2000 (as determined by tipping bucket rain gage, equipped with a heater to melt the snow as fast as it falls) averaged about 0.8 mm. hr.$^{-1}$ For crystal snow $Z=500\ R^{1.6}$ [9], which for a Z of 234 mm.^{6} m.$^{-3}$ gives $R=0.6$ mm. hr.$^{-1}$ Although the actual snow at the ground appeared to consist generally of aggregates, there may have been more in the form of crystals at 0.6 km. where the lowest reflectivity measurements were made.

The reflectivity at 5–6 km. was relatively high at the beginning of the period when the winds were from the WSW, but with the coming of the westerlies the reflectivity drops markedly: by more than 10 db. between 1400 and 1530 EST. The reflectivity increases at these levels by about 5 db. between 1550 and 1630 and remains relatively constant to 2000 EST. The increased reflectivity at 1630 may be related to the approach of the mesoscale wave; however, the lack of data between 1630 and 1800 EST prevents any detailed analysis.

It is noteworthy that, although the horizontal variability in reflectivity from 1330 to 1550 EST was about 12 db. at 5 to 6 km., it was only about 4 db. at elevations up to 4 km. This reduction in variability at lower levels is evidently the result of the generation of precipitation of variable intensity aloft and its blending together at lower levels due to particle sorting in streamers.

7. FIELD OF DIVERGENCE AND VERTICAL VELOCITY

In figure 5 positive values represent divergence and negative values convergence. During the first 3 hr. of data, there are five layers alternating between convergence and divergence. A weak convergence is indicated in the lower easterlies and then weak divergence extends through the transition layer. The moderate convergence in the southwesterlies (2.5–3.5 mm.) is replaced by strong divergence in the layer 3.5 to 5 km. where the wind is veering with height from southwest to west. Strong convergence occurs above in the layer 5.5–6.4 km. where the wind backs with height from W to SW. Finally, strong divergence is located near the top of the cloud where the wind direction remains constant with elevation. In the last 2 hr. there is weak divergence in the easterlies and transition layer and weak convergence in the southwesterlies near 3 km., but the layer of strong divergence in the westerlies has essentially disappeared, although there is some indication of its reestablishment at 3 km. at 1950 EST. Moderate convergence occurs near 5 km. again where the westerlies back with height to the SW. Very strong divergence is indicated near the top of the storm.

Figure 6 shows the field of vertical velocity with respect to datum level at 0.74 km. as derived from figure 5 using equations (3). A indicated previously, the chart is subject to error due to the lack of accurate divergence values at low levels. For this reason the individual values have been omitted.

At the beginning of the period in figure 6, updrafts prevail at all levels with strongest values of 30 cm. sec.$^{-1}$ near 6 km. This was evidently associated with the relatively high reflectivity observed at upper levels near the beginning of the period. After 1400 EST downdrafts occur in the westerlies at 4–5 km. with only weak updrafts above, and this pattern is in agreement with the sharp reduction in reflectivity that occurs during the period. Updrafts near 30 cm. sec.$^{-1}$ are located at 6 to 7 km. in the upper southwesterlies, but strong subsidence is indicated above 6 km. from 1800–1900 EST. The curious alternation of updraft and downdraft near 4 km. after 1530 EST may be in error, since it is difficult to explain and no indication of this feature exists in the reflectivity pattern. The strong upward motion of 40 cm. sec.$^{-1}$ centered at 6.7 km. at 1930 EST is accompanied by only a slight increase in reflectivity. However, updrafts of this magnitude generally produce very localized concentrations of reflectivity and the records at the time do show some sharp changes in reflectivity with azimuth.

8. HORIZONTAL CROSS-WIND SHEAR

Figure 7 shows the direction of increasing wind speed in cross-wind shear, as determined from the spacing of the upwind and downwind Doppler velocity maxima. A reading is shown only for those cases where the spacing between successive maxima (upwind and downwind) differed from 180° by 10° or more. In the lower northeasterly current at 0.75 km. there is a tendency for winds to be stronger to the left of the wind or toward the SSE. In the southwesterlies at 3 km., higher speeds are to the right of the wind or again towards the SE. This is generally in agreement with rawin observations for 1900 EST March 15, 1967. At 5.5 km. (near the 550-mb. level) in the westerlies, there appears to be a tendency for winds to be stronger to the north, from 1430–1630 EST; after 1900 EST stronger winds are indicated first toward the SE then toward the NW. The 500-mb. synoptic map at 1900 EST shows little or no gradient between Albany and New York but stronger winds were observed at Nantucket than at Portland. At 7.1 km. (400 mb.) in the southwesterlies stronger winds to the SE are indicated at 1900 EST by the radar data in agreement with the observed winds at that time. In general, therefore, the method appears to provide a qualitative estimate of horizontal

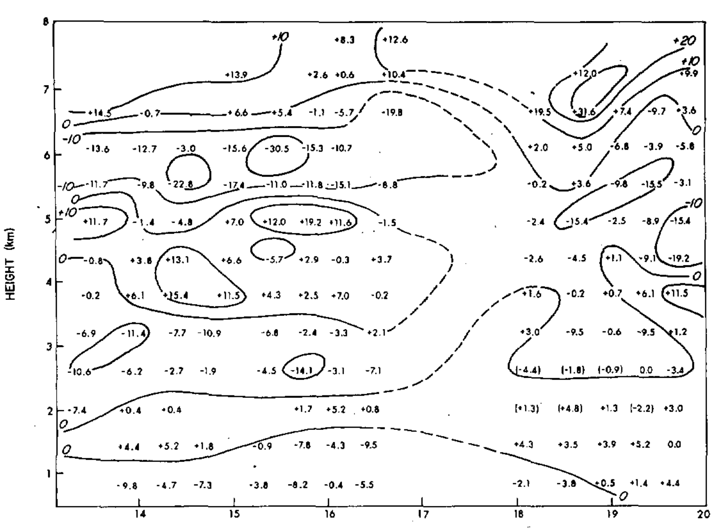

FIGURE 5.—Divergence (+) and convergence (−) in 10^{-5} sec.$^{-1}$, March 15, 1967.

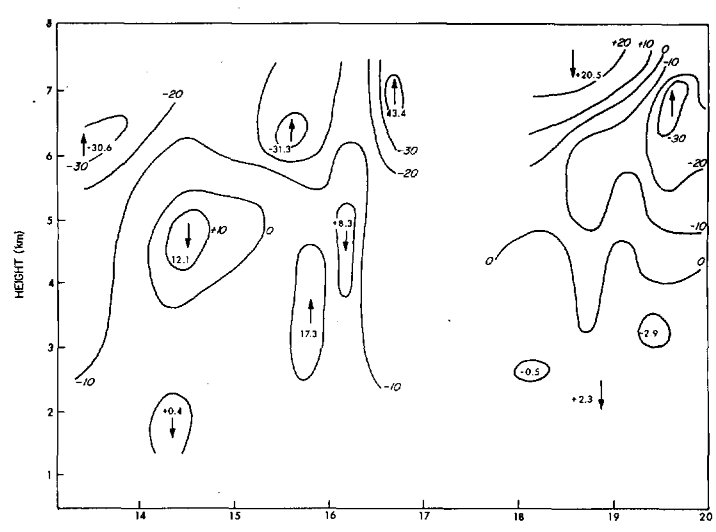

FIGURE 6.—Vertical velocity (cm.sec.$^{-1}$), March 15, 1967.

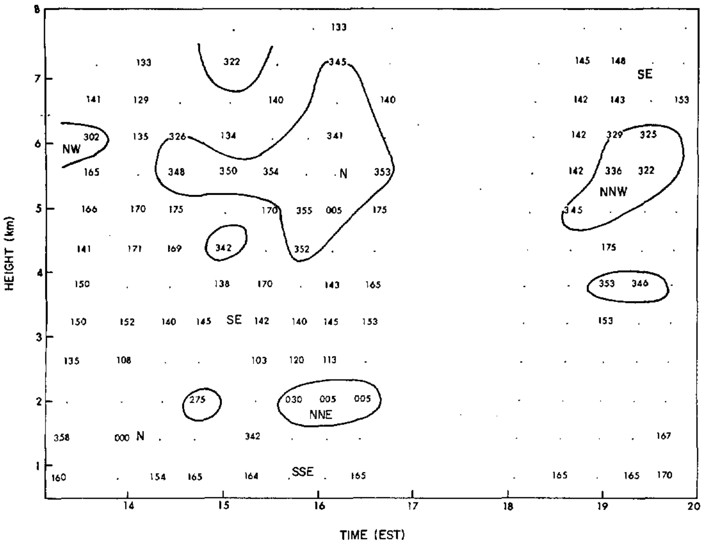

FIGURE 7.—Direction of increasing wind speed in cross-wind shear, March 15, 1967.

wind shear which is in agreement with the shear determined from the synoptic maps.

9. CONCLUSION

The Doppler radar data from the storm of March 15, 1967, and data from other storms indicate that the wind field in continuous precipitation may be quite variable over periods of about 1 hr. This variability is undoubtedly associated with the variability in the generation of precipitation aloft. Although on the synoptic scale an apparently straight flow pattern occurs ahead of a major trough, more detailed examination of the wind structure often indicates the existence of mesoscale short waves of relatively low amplitude imbedded in the flow pattern ahead of the trough. These minor short waves are evidently responsible for a banded structure in the generation of the precipitation. With wind shear and particle sorting, the bands of heavier precipitation fall in sloping trails that blend together at low levels. This causes the precipitation intensity at low levels to be relatively uniform compared to that at high levels. In the March 15, 1967, storm the reflectivity aloft changes by more than 12 db. within 1 hr. while at low levels it changes by only 4 db.

Generally in precipitation in the advance of a cyclonic storm approaching New England from the south, the wind veers with height from northeasterly at low levels to southwesterly at 1.5–3 km. and then veers gradually with height to a westerly direction with increasing speed. The March 15, 1967, storm is unusual in that a westerly current at 4 to 5 km. was sandwiched in between a southwesterly flow. The passage of minor short-wave troughs was indicated by the changes of wind near 5 km. from WSW to W, back to SW, and returning to W all within a 6-hr. period.

The increase in depth of the NE to E current above the surface was evidently responsible for maintaining the precipitation as snow. Between the easterly current and the upper westerly flow the layer of southwesterlies decreased in thickness and increased in speed, probably as a result of the increasing depth of the lower easterlies and the descent of the upper westerlies.

The pattern of divergence for the first 3 hr. had five relatively thin layers, 1 km. or less, of alternating divergence and convergence. A somewhat similar pattern was found by Harrold [1], but in the case studied by Browning and Wexler [3] there were only three layers with the depth of the central convergent layer about 3 km. thick. The layers of convergence and divergence in the March 15, 1967, case appeared to be associated with the wind pattern: strong divergence where the wind veered from SW to W, strong convergence where it backed from W to SW and then divergence in the SW flow near the top of the storm. The maximum values of the divergence, about $\pm 2\times 10^{-4}$ sec.$^{-1}$, are in agreement with those computed by Caton [2] and Browning and Wexler [3]. The values of Harrold [1], which are an order of magnitude larger, may be in error.

The computed field of vertical velocity is at a disadvantage because of the compounded errors aloft due to poor data in the region of light winds at 1.5–2 km. Nevertheless, certain features in the overall pattern appear reliable: little or no vertical motions up to about 4 km.; some descending motion at 4–5 km. during the period when the westerlies were being established, and relatively strong updrafts of 15–30 cm. sec.$^{-1}$ just above the westerlies. The magnitude of the vertical motions are in agreement with those found by Caton [2] and Browning and Wexler [3] in continuous precipitation.

The techniques used in the March 15, 1967, storm and in previous storms provide the first systematic and reasonably accurate method to extend mesoscale meteorology up from the ground—at least as far as winds are concerned. The results indicate that mesoscale features are important not only in convective precipitation but also in continuous precipitation.

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